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A Micro-Meteorological mission for global network science on Mars: rationale and measurement requirements

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Abstract. The general circulation of the Martian atmosphere and how it relates to the planet's climate system are the main goals of post-Viking atmospheric science investigations. An unambiguous determination of the wind systems associated with the general circulation requires simultaneous measurements from at least 15–20 globally-distributed surface stations and a single, high inclination, short period orbiter. The required number of stations would be too costly using conventional multi-instrumented lander designs. However, truly global network science for meteorology can be accomplished using landers instrumented for a single measurement, namely pressure. Pressure is the most important meteorological parameter and pressure sensors do not require deployment or orientation. This facilitates the design of relatively simple landers that are sufficiently small and light that the required number can be launched on a single Med-Lite launch vehicle (see the accompanying paper by Merrihew *et al.* (*Planet. Space Sci.* **44**, 1385–1393, 1996)). Of equal importance, the low power demand of the stations would allow them to last for a Martian year and provide the necessary seasonal coverage. The rationale and measurement requirements are given for such a network, and it is shown how it would contribute to the understanding of the Martian general circulation and climate system. In particular, the number and siting of stations is discussed, and the basic science specifications for barometric pressure sensors and an atmospheric sounder. The proposed mission would, for the first time, observationally define the seasonally-varying global wind field on another planet. Published by Elsevier Science Ltd

1. Introduction

The highest priority atmospheric science objectives in the post-Viking era of Mars exploration concern the nature of the planet's general circulation. Briefly, the general circulation consists of global-scale wind systems which can vary significantly in space and time. Typically, these systems are defined in terms of their time and zonally averaged structures (zonal mean circulation), and departures from them (eddies) (e.g. Lorenz, 1967). The time averaging interval is generally taken to be long enough to include many daily cycles, but sufficiently short to detect seasonal trends. For Mars, the general circulation interacts with the surface, transports dust and volatiles around the planet, and influences the CO_2 cycle. The latter is a large and unique component of the planet's climate system. While the spacecraft missions of the 1960s and 1970s have identified some aspects of the general circulation, they lacked the temporal and spatial coverage needed for its full characterization. Future missions will address this issue, but those currently planned will not have the type of coordinated observations needed to calculate wind fields: namely, simultaneous measurements from orbit and a large network of landers. Therefore, no currently planned missions can achieve the basic objective of measuring what is defined as the general circulation.

In this paper, we set down the science rationale and measurement requirements for a mission that would provide a large network of landers. We call this the Micro-Meteorology (or "μ-Met") mission.[1] This is a Discovery class mission that would launch 16 probes to Mars aboard a Med-Lite launch vehicle. The probes would be deployed on approach from a spin-stabilized carrier using a single propulsive time-of-arrival adjustment to achieve global coverage. The entry, descent, and landing of each probe would utilize an ablatable aeroshell, a parachute, and crushable material to reduce the impact shock. On the surface, the battery-powered stations would measure pressure, and pressure alone 25 times a day for 1 Martian year (1.9 Earth years). The data would be relayed to an orbiter

[1] The term "μ-Met" is meant here to emphasize the small size and highly-focused nature of the landers—not the branch of meteorology that deals with small-scale phenomena.

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through a UHF link and then transmitted to Earth. Details on the engineering aspects of the mission are given in a companion paper by Merrihew *et al.* (1996).

The immediate objective of the μ-Met mission is to map the global-scale surface pressure field over time. Given the Viking experience, such data alone can reveal a great deal of information about Mars meteorology. However, the μ-Met network would not be a stand-alone mission. For atmospheric science, simultaneous measurements from the surface and orbit should be the focus of missions in the next century. Thus, we advocate that the μ-Met mission be flown in conjunction with an orbiter and that the orbiter function not only as a data relay, but as an observing platform as well. The most favorable orbiter would be that proposed for InterMarsNet (in 2003) which is baselined to include an atmospheric sounder. With simultaneous measurements from the surface and orbit it is possible to determine the horizontal wind field and, in turn, the time-varying three-dimensional structure of the general circulation.

We describe how global surface measurements of a single parameter, when combined with remote sensing measurements from orbit, can satisfy the science objectives related to the general circulation of the Martian atmosphere. Previously, such a strategy has been suggested at a workshop on Martian network meteorology held at NASA/Ames Research Center (Haberle, 1990) and has also been advocated by Leovy (1990). We draw heavily from these works and build on them by providing a more quantitative description of the rationale and measurement requirements. We begin with a brief discussion of why Mars is important and what we know about its general circulation and climate system. We then present related science objectives. These arise from the absence of data to systematically characterize the Martian atmospheric circulation. The rest of this paper discusses how the general circulation can be measured, how currently planned missions address those measurements, and how the μ-Met concept fits within the context of those plans.

2. Background

2.1. *Why is Mars important?*

Mars is an important target for atmospheric science investigations because it is the most Earth-like planet in the solar system. Table 1 gives a comparison of the important parameters. Its thin, nearly transparent CO_2 atmosphere allows sunlight to reach the surface where it is absorbed and re-radiated in the infrared. Thus, as is the case on Earth, the Martian atmosphere is heated mainly from the surface. Furthermore, its spin axis inclination and rotation rate are comparable to Earth's which produces similarity in large-scale dynamics and seasonal change. What makes Mars unique, however, is the very short thermal response time of its atmosphere ($\sim$2–4 days), the fact that it has no oceans to drive a complex hydrological cycle, and its large orbital eccentricity. The latter leads to a 40% variation in insolation during the course of a Martian year. Even more exceptional is the $\sim$25% seasonal fluctuation in atmospheric mass due to the condensation and sublimation of the major constituent, and the occasional development of global-scale dust storms which can dramatically alter the thermal drive and structure of the general circulation. The similarities and differences with the Earth's climate system make Mars an ideal natural laboratory for studying the general circulation of rapidly rotating, differentially heated atmospheres.

Table 1. A comparison of parameters for Mars and the Earth

	Mars	Earth
Planetary properties		
Mass, kg	6.46×10^{23}	5.98×10^{24}
Radius, km	3394	6378
Gravity, m s^{-2}	3.72	9.81
Rotation rate, 10^{-4} s^{-1}	0.7088	0.7294
Orbit eccentricity	0.093	0.017
Obliquity	25.2°	23.5°
Year length, Earth days	687	365
Solar day lengths, s	88,775	86,400
Solar constant, W m^{-2}	591	1373
Effective temperature, K	210	256
Atmospheric properties		
Principal constituents	CO_2 (95.3%) N_2 (2.7%) ^{40}Ar (1.6%) O_2 (0.13%)	N_2 (78.1%) O_2 (20.9%) ^{40}Ar (0.9%) CO_2 (0.03%)
Mean surface pressure, mbar	6.1	1013
Mean surface temperature, K	218	288
Mean molar mass, g	43.4	29
Mean scale height, km	10.8	7.5
Adiabatic lapse rate, K km^{-1}	4.5	9.8
Mean lapse rate, K km^{-1}	$\sim$2.5	6.5
Richardson number	$\sim$30	$\sim$65
Thermal Rossby number	$\sim$0.2	$\sim$0.03
Rossby radius, km	920	1150
Radiative time constant, days	1–4	>20
Hadley cell overturning time, days	$\sim$50	$\sim$200

2.2. *What do we know about the general circulation of the Martian atmosphere?*

From the spacecraft missions of the 1960s and 1970s several components of the Martian general circulation have been identified (see the review by Zurek *et al.* (1992) and references therein). These include the net meridional mass flow due to the condensation and sublimation of CO_2 in the polar regions, the seasonally-varying mean meridional circulation, traveling baroclinic weather systems in the northern hemisphere, thermal tides, and smaller scale phenomena such as regional slope winds and internal gravity waves. Yet the available data lack the temporal and spatial coverage needed to fully characterize these systems. Consequently, we do not know their full three-dimensional structure or how they vary with season and dust loading. Instead, we have had to rely on atmospheric models, and general circulation models (GCMs) in particular. The first such model was developed by Conway Leovy and Yale Mintz in 1969, and later by Jim Pollack

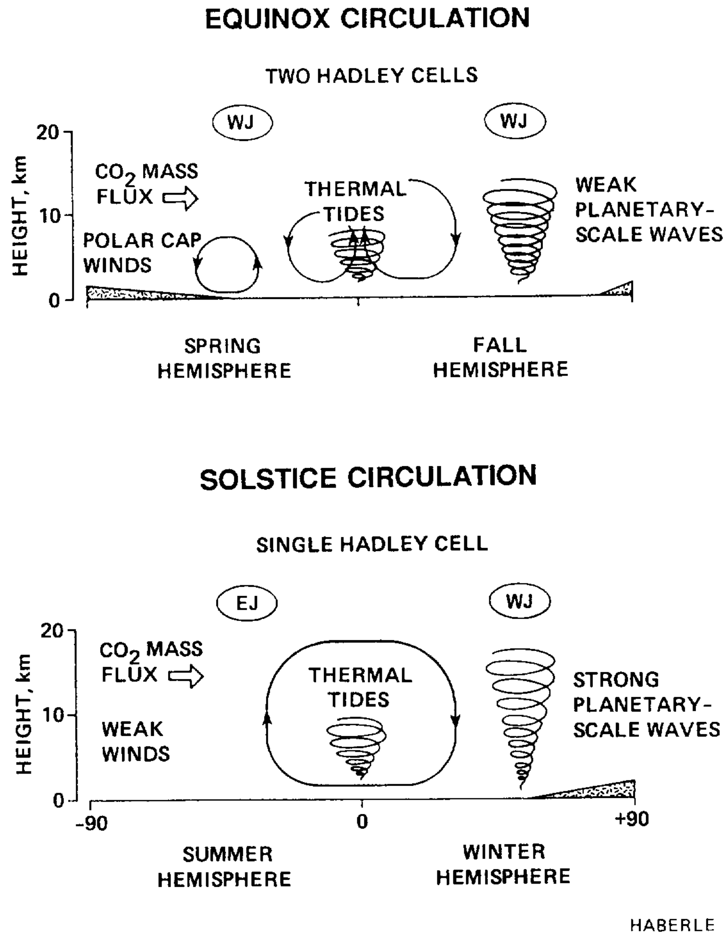

Fig. 1. A schematic diagram showing the main components of the general circulation on Mars. WJ means westerly jet, EJ means easterly jet. Taken from Jakosky and Haberle (1992)

and collaborators at NASA/Ames Research Center. Since then a number of groups have been developing Mars GCMs in anticipation of the acquisition of global synoptic data from forthcoming missions. These groups include NASA's Goddard Institute for Space Studies; the Geophysical Fluid Dynamics Laboratory (GFDL) at Princeton; the Oxford/Reading group in the U.K.; and a group at the Laboratoire de Météorologie Dynamique (LMD) in Paris. Yet while the models developed by these groups have been useful in illuminating aspects of the Martian general circulation, they need validation if we are to believe their predictions. A schematic illustration of the various components of the Martian general circulation based on these models and the limited available data is given in Fig. 1.

2.2.1. *The Hadley circulation.* According to the models, the Martian general circulation is dominated by the zonal-mean circulation (Haberle *et al.*, 1993). This component of the circulation is most prominent at the solstices when it consists of a single cross-equatorial Hadley circulation, tropical easterlies, and strong low-level westerly jets in the summer subtropics. Because the Martian atmosphere is thin and radiatively active, heating rates (per unit mass) can be quite substantial ($\sim$ few $\times 10\,\mathrm{K\,sol^{-1}}$, where $1\,\mathrm{sol} = 24.66\,\mathrm{h}$ is a Martian solar day). Consequently, compared to Earth the overturning time scale associated with the Martian Hadley circulation is short, and the mean equator-to-pole temperature gradient is large. The latter gives a larger thermal Rossby number for Mars so that its midlatitude jet streams are stronger than those on Earth. Yet there are several aspects of the simulated zonal-mean circulation we would like to verify. According to the models, the Martian Hadley circulation at northern winter solstice is about twice as intense as that at southern winter solstice, a variation greater than can be explained by solar forcing alone (Joshi *et al.*, 1995a). The models also indicate that dust intensifies and expands the Hadley circulation, both vertically and latitudinally, because it absorbs solar radiation (Haberle *et al.*, 1993). We would like to know if these simulated changes are correct.

In some respects, we already know that certain aspects of the model simulations are incorrect. A major shortcoming, for example, is the inability of models to simulate the warming of the north polar upper atmosphere observed during the second global dust storm of 1977 (Martin and Kieffer, 1979). At the time the warming occurred, the surface pressure at the Viking Lander 2 (VL-2) site (48°N) increased, winds shifted from westerly to northeasterly, and transient eddy activity weakened. These changes suggest that the descending branch of the Hadley circulation shifted poleward of the lander site (Haberle *et al.*, 1982). While the models do show poleward movement, it is not enough to explain the observations (Murphy *et al.*, 1995). Consequently, the upper polar atmosphere remains cool in the models. This deficiency suggests that the models are not capturing some fundamental dynamical process. But without more detailed observations, it is difficult to determine what these processes might be. Thus, we would like to know better how the thermal structure and surface pressure fields change during these events. For example, over what depth does the warming occur? Do the transient eddies truly weaken, or are their storm tracks shifted poleward? And is there evidence at other longitudes for an expanded Hadley circulation?

Though it is common to think of the Hadley circulation as being more or less zonally symmetric, the models tell us that it is not. In particular, the low level flow associated with the Hadley circulation has been found to be very longitudinally asymmetric (Joshi *et al.*, 1994). Instead of a uniform cross-equatorial flow, much of the meridional motion appears to be concentrated into intense (40–50 m s^{-1}) low level jets that form along the eastward flanks of the Tharsis plateau and Syrtis Major. The existence of these jets appears to be the Hadley cell's response to the presence of large-scale topography. Similar phenomena, known as Western Boundary Currents, occur in the Earth's atmosphere and oceans. But do such currents really exist on Mars, and if so, do they form at the longitudes predicted by the models? How do they vary with season, and what role do they play in the exchange of dust and water between hemispheres?

2.2.2. *Transient eddies.* Transient baroclinic eddies were detected by the Viking landers, or VLs (e.g. Tillman *et al.*, 1979; Barnes, 1980), and have been seen in model simulations since the pioneering work of Leovy and Mintz (1969). They are believed to be manifestations of an unstable winter midlatitude westerly jet stream. This jet is predicted by the models to be very strong (100–200 m s^{-1}) and very deep (>40 km). We would like to know if indeed there are westerly jet streams on Mars as deep and intense as predicted by the models, and if so, at what point do they close off? This kind of information would allow us to assess the kinds of dynamical processes that characterize transient eddies.

One particularly surprising result of the models is the difference in behavior of midlatitude systems between the northern and southern hemispheres. According to the models, these wintertime disturbances are much more vigorous in the northern hemisphere than in the southern hemisphere (Barnes *et al.*, 1993) and it appears that the different slopes between the two hemispheres contribute to the asymmetry. Weaker solar forcing during southern winter also plays a role. But even the existence of baroclinic eddies in the southern hemisphere is not known. Thus, we would like to determine what differences there are in the structure and amplitude of midlatitude baroclinic waves in the northern and southern hemispheres. If there are significant contrasts, as the models suggest, then it is plausible that the asymmetry in the current H_2O cycle (see Jakosky and Haberle, 1992) is linked to them.

Another surprise predicted by the models is that winter transient eddies appear to favor certain regions for growth and decay, at least in the northern hemisphere. Time-filtered temporal variance and co-variance fields of wind and temperature taken from the NASA/Ames GCM show distinct maxima in the midlatitude low-relief regions of Arcadia, Acidalia, and Utopia (Hollingsworth *et al.*, 1996). These "storm zones" appear to be organized by Mars' continental-scale topography rather than thermal contrasts associated with spatial variations in albedo and thermal inertia. But do such storm zones exist? And what role, if any, do they play in the transport of momentum, heat, moisture, and dust? The Ames model suggests a considerable role in this regard, but we need better observations to validate their existence.

An intriguing aspect of transient baroclinic eddies on Mars is their occasional regularity. Time series analyses of Viking lander meteorological data show that most of the synoptic variance is accounted for by several dominant periodicities (Barnes, 1980). On various occasions (fall of the first Viking year, winter of the second year), many wave cycles were observed with little change in their amplitude and period. At other times, the waves were less regular. This occasional high degree of regularity is very different from the Earth's irregular weather systems and we would like to know why. Recent work by Collins *et al.* (1995, 1996) point to a significant role for diurnal forcing. In model simulations using daily-averaged forcing they find an unrealistically high degree of repeatability throughout the fall, winter, and spring seasons, a result at least partly attributable to the short radiative time constant of the Martian atmosphere. With diurnal forcing included, however, they find a wave behavior that more closely resembles that seen by the Viking landers. Evidently, the diurnal cycle facilitates transitions in the dominant wave number, transitions which are also seen in the data. But the connection between diurnal forcing, which is relatively weak at winter midlatitudes, and wave response is unclear. This connection could be elucidated with a network of landers in the tropics and at midlatitudes.

2.2.3. *Stationary waves.* Given the presence of planetary-scale variations in topographical relief that are on the order of a scale height, we expect that stationary waves should be a prominent feature of the general circulation of Mars. Yet the available data are ambiguous (Conrath, 1981; Jakosky and Martin, 1987; Banfield *et al.*, 1996), so again, we have had to rely on models. The models suggest that very deep stationary eddies develop in midlatitudes during winter with zonal wave number 1 dominant in the southern hemisphere and wave number 2 dominant in the northern hemisphere (Barnes *et al.*, 1996). This asymmetry appears to be due to the prominence of wave number 1 in southern hemisphere topography since the simulated mean zonal flows are similar during southern and northern winter. Is this the case, or is there a different stationary wave number pattern? Are the zonal flows of the two hemispheres really that similar? And is wave number 1 really prominent in southern hemisphere topography? Current topographic data sets have uncertainties of a kilometer or so, and have very little information for the higher latitudes. Clearly, this is an important boundary condition for general circulation studies.

We would also like to know to what extent the stationary waves are mechanically vs. thermally forced. The latter arises because elevated regions tend to act as heat sources on Mars because of the short thermal response time of its atmosphere. The models suggest significant roles for both mechanisms, with mechanical forcing important at high latitudes and thermal forcing important at low latitudes (Webster, 1977; Pollack *et al.*, 1981; Hollingsworth and Barnes, 1996; Barnes *et al.*, 1996). Since each type of forcing produces a unique response, we could distinguish between them with a long-term global synoptic data set of winds and temperatures.

Simulated quasi-stationary waves increase in amplitude with atmospheric dust loading and in the northern hemisphere, the dominant wave number decreases from 2 to 1. This is particularly evident in the recent interactive dust transport simulations of Murphy *et al.* (1995) which show height perturbations of isobaric surfaces at upper levels (0.3 mbar) as large as 4 km. These changes are brought about by changes in the zonal-mean state and/or a wave number 1 resonance. Changes in wave forcing do not appear to be important in this regard. Similarly, Hollingsworth and Barnes (1996) found a resonant wave number 1 and far weaker wave number 2 in linear stationary wave calculations for dusty Mars conditions. These results are intriguing but since the models yield only a modest polar warming, they are probably not correctly simulating the real changes to the stationary wave pattern. What are these changes and what role, if any, do stationary waves play in the polar warming phenomenon?

2.2.4. *Thermal tides.* The large diurnal temperature variations on Mars drive powerful thermal tides that can generate a 10% daily fluctuation in surface pressure (Leovy, 1981). Both theory and models predict significant topographic modulation of the thermal tides (e.g. Zurek, 1976; Wilson and Hamilton, 1996). The presence of zonally varying topography and surface properties on Mars excites the eastward propagating Kelvin wave components of the tides. Of particular interest is the normal mode wave number 1 component which has a period close to 1 sol and which may therefore be resonantly excited by the daily heating cycle (Hamilton and Garcia, 1986; Zurek, 1988). The interaction of the diurnal sun-following (westward) and Kelvin (eastward) tides is expected to reinforce in some longitudes and cancel in others. This phenomenon may play an important role in the onset of global scale dust storms (Leovy and Zurek, 1979; Tillman, 1988). A properly configured low latitude network of landers could distinguish various tidal components and determine the longitudes of reinforcement and cancellation, if they exist.

Also, in Wilson and Hamilton's model, the amplitudes and phases of the diurnal tide were found to be in good agreement with Viking observations, except during northern summer when the simulated amplitudes were larger by a factor of two. They found that much of the amplitude of the diurnal tide was attributable to the wave number 1 Kelvin mode and this could be weakened in their model if atmospheric dust was concentrated over upland regions. We would like to know if the circulation does indeed organize itself in such a way as to concentrate dust over the highlands. If so, it has implications for the accumulation of fine material in these regions (Christensen, 1986).

2.2.5. *Small-scale phenomena.* Though the main focus here is on the general circulation, there are two important small-scale phenomena we mention for completeness. These are boundary layer turbulence and breaking gravity waves. Boundary layer turbulence determines the magnitude of surface fluxes and is therefore an important forcing function. During the day, the boundary layer can extend throughout the lowest scale height thereby communicating with a major fraction of the atmosphere. Though some useful information has been gleaned from the limited data available (Sutton *et al.*, 1978; Haberle *et al.*, 1993; Tillman *et al.*, 1994), much remains unknown. Basic similarity theory seems to be applicable to the surface layer of Mars, but the kind of measurements to verify it (high vertical resolution, near-surface profiles of wind and temperature) are lacking. And we really have no good information on the structure of the mixed layer above it. Boundary layer studies, therefore, require much different landers than those advocated here. The InterMarsNet landers are, in principle, much better suited for such measurements.

Internal gravity waves are most commonly generated by flow over topography. In the absence of critical levels they can propagate to great heights and break (Barnes, 1990; Joshi *et al.*, 1995b). In doing so they deposit their momentum and mix the local environment. Breaking gravity waves play an important role in the Earth's middle atmosphere momentum balance, and may play a similar role on Mars. Furthermore, because the breaking level is a sensitive function of the mean flow, there are potential feedbacks that may be important (Joshi *et al.*, 1995b; Murphy *et al.*, 1995). We will not be able to detect breaking gravity waves directly with the kind of coordinated measurements proposed here. However we can (in principle) use these measurements to diagnose the effects of breaking gravity waves on the zonal mean circulation (see Section 3.1).

2.3. *How is the general circulation linked to the climate system?*

The present Martian climate system consists of three main components that involve the cycling of dust, water, and carbon dioxide between various atmospheric and non-atmospheric reservoirs. The general circulation is related to these cycles through its influence on the surface fluxes of dust and water, their transport around the planet, and on the heat balance of the polar regions.

2.3.1. *The dust cycle.* The most important element of the present-day climate system is dust. Its existence in the atmosphere alters the heating rates and therefore the thermal drive and intensity of the circulation itself. This changes the coupling between the surface and atmosphere, the transport of material around the planet, and the heat advected into the polar regions. Dust is therefore a key parameter influencing the Martian climate system.

The most pressing issue regarding the dust cycle concerns the mechanisms responsible for the origin and decay of planetwide dust storms. These quasiannual events lift dust from the surface and spread it over virtually the entire planet. What triggers these spectacular events? While many theories have been suggested (see the review of Zurek *et al.* (1992)), none completely accounts for all the observations. And what is particularly puzzling is the

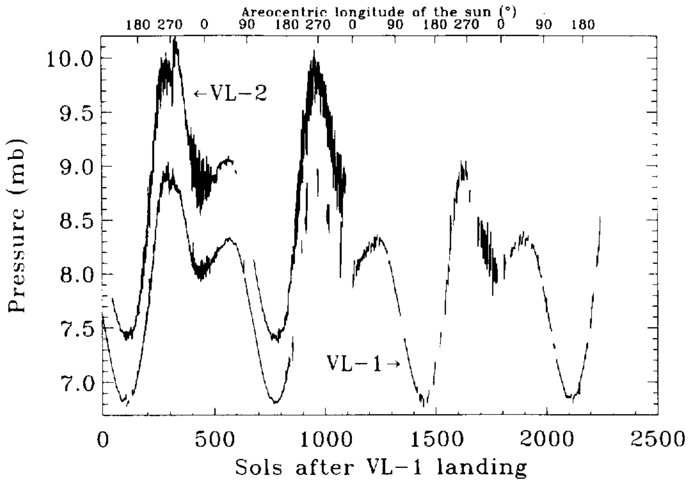

Fig. 2. Sol-mean pressure at the two Viking lander sites for 3.3 Mars years. Viking Lander 2 (VL-2) pressures are generally greater than those at Viking Lander 1 (VL-1) due to its lower elevation. Gaps in the data are mostly because of the use of the Deep Space Network for Voyager encounters with Jupiter and Saturn. The top horizontal axis shows the aerocentric longitude (often denoted by as L_s). 0° corresponds to the northern spring equinox, 90° to the northern summer solstice, and so on

fact that these storms do not occur every Martian year as was previously believed (Zurek and Martin, 1993). Thus, theory must also explain their interannual variability.

Certainly, feedbacks between dust heating and wind systems must play and important role. The questions are which wind systems, and in what way? The recent modeling work of Murphy *et al.* (1995) has shown that positive feedbacks between dust heating, the Hadley circulation, and thermally-forced tides can increase surface stresses in low southern latitudes enough to cause significant lifting of dust in those regions. This result supports the global dust storm model proposed by Leovy *et al.* (1973). But does this really happen? Alternatively, are there other mechanisms such as free-mode triggering (Tillman, 1988), Hadley cell expansion (Schneider, 1983), or low-level jets (Joshi *et al.*, 1995a) that are just as important? Furthermore, by what mechanism do these storms decay? The Murphy *et al.*, calculations show no negative feedbacks in the system; though the atmosphere does stabilize (making it difficult to mix dust vertically), surface stresses monotonically increase with dust loading. Does this mean the source of dust is limited, or are the models missing something? We need better observations of dust storm wind systems to answer these questions.

2.3.2. *The CO_2 cycle.* The signature of the CO_2 cycle was measured by the Viking landers and is shown in Fig. 2. The semi-annual oscillation of daily-averaged surface pressures largely reflects the sublimation and condensation of CO_2 in the polar regions as predicted by Leighton and Murray (1966). The deepest minimum in these data occurs during the long southern winter when the planet is furthest from the sun. However, the difference between the northern and southern winter minima is more pronounced in the data than that of a simple Leighton/Murray type model and can only be reconciled if additional heat sources are included in the north cap's winter energy budget (see James *et al.* (1992) for a recent review). These extra heat sources can take the form of a lower cap emissivity (James and North, 1982), a higher conducted heat flux from the ground (Wood and Paige, 1992), or increased atmospheric emission due to dynamical heat transport (Pollack *et al.*, 1990, 1993).

A difficulty in determining the relative contribution of these factors is the uncertainty in the *actual* global atmospheric *mass* budget. Until recently, it was generally assumed that the curves in Fig. 2 represented that mass budget. However, a number of groups have shown that basic meteorological effects can significantly modulate the CO_2 signal (Talagrand *et al.*, 1991; Pollack *et al.*, 1993; Hourdin *et al.*, 1995). This arises from the redistribution of atmospheric mass by winds, and from changes in the atmospheric scale height associated with the seasonal temperature wave. As a consequence, the surface pressure measured at one site may not have the same seasonal variation as that at another site. Indeed, a careful examination of the Viking Lander pressure data supports this conclusion (Pollack *et al.*, 1993; Hourdin *et al.*, 1995). Thus, we still do not know accurately how much CO_2 actually condenses in the polar regions.

2.3.3. *The water cycle.* The amount and variability of water vapor in the Martian atmosphere provide clues about the nature and distribution of water in the surface. At present, the only positively identified non-atmospheric reservoir for water is surface ice in the north polar region (Farmer *et al.*, 1976). By late spring, the CO_2 ice that condensed during the previous winter completely disappears. This exposes an underlying residual water ice cap which then sublimes.

A major question regarding the present water cycle concerns the fate of this sublimed water. Is it transported to lower latitudes and other reservoirs? Candidate reservoirs include the global regolith and the south polar CO_2 ice deposit which is able to survive the summer season unlike its northern counterpart. The Viking Mars Atmospheric Water Detector (MAWD) data and modeling studies hint that transport to lower latitudes and eventual incorporation into the regolith does occur (Jakosky and Farmer, 1982; Haberle and Jakosky, 1990). If CO_2 ice does survive at the south pole all year long, as was the case during the first Viking year, then it will act as a cold trap for any water vapor that comes in contact with it. What we would like to know, therefore, is the annual mass balance of the northern residual water ice cap.

There are, of course, three possibilities: the cap is either gaining water, losing water, or there is no net change. Arguments in support of each of these possibilities have been advanced. Two components of the general circulation, the condensation flow and the Hadley circulation, have the potential to act as south-to-north pumps and therefore build up the cap. The condensation flow is seasonally and hemispherically asymmetric and is strongest in the southern hemisphere during spring when it is directed northward. On decadel time scales, advection of water by this asymmetric condensation flow can transfer water from south to north (James, 1985). The Hadley circulation may have the same effect since its thermal structure during northern summer, when the planet is near aphelion, tends to concentrate water in its lower (northward) branch (Clancy *et al.*, 1996). There is less of a tendency to concentrate water at low altitudes during southern summer since the planet is near perihelion at this season and the atmosphere is warmer at upper levels. Consequently, the yearly averaged Hadley cell transport would be toward the north. Of course, both these effects result from the planet's eccentricity. On the other hand, Jakosky (1983) has argued that atmospheric circulations tend to mix water down gradient and that therefore the north cap must be loosing water to the south cap during the present epoch. While the recent modeling work of Houben *et al.* (1996) supports this conclusion, it also demonstrates that the regolith can be such an effective buffer, that the net exchange is very small. Davies (1981) arrived at a similar conclusion but invoked dust storms rather than regolith buffering as a mechanism to stabilize the system. Thus, we have three, mutually exclusive possibilities that cannot be reconciled with the available data. To distinguish among them requires much better observations of the general circulation, the vertical distribution of atmospheric water vapor and ices, and the nature of Martian surface materials.

2.3.4. *Climate change and atmospheric evolution.* Several lines of evidence suggest that Mars has experienced significant climatic change and that its atmosphere has evolved over time. Evidence for recent climate change (past tens of million years) can be found in the polar regions which are characterized by extensive layered deposits. These appear to result from long-term modulation of the transport and deposition of dust and water vapor (see Kieffer and Zent (1992) for a review). The modulation is driven by changes in Mars' orbital parameters which are quite substantial compared to the Earth's. The planet's obliquity, for example, is now believed to vary chaotically on time scales of 10^7 years and may have reached values as high as 60° (Touma and Wisdom, 1993; Laskar and Robutel, 1993). Variations in the orbital parameters will alter the seasonal and latitudinal distribution of sunlight and, therefore, the general circulation. This, in turn, alters the transport and deposition of dust and volatiles.

It is quite probable that in the foreseeable future, major advances in our understanding of this kind of climate change will come from studies of the current general circulation and climate system. Such studies can be used to validate and improve the models which can then be used to explore past climates. But only after we have some confidence in our predictions for the current system—the one we can directly observe—will we be able to take this step.

There is also evidence for climate change on much longer time scales. But for these earlier epochs, studies of the current climate system are perhaps less relevant. This is mainly because early atmospheres may have involved significant mass and/or compositional differences from the present atmosphere (e.g. Haberle *et al.*, 1994). For example, the fluvial features on the flanks of Alba Patera (Gulick and Baker, 1990) and the suite of putative glacial landforms in the southern hemisphere (Kargel and Strom, 1992) may have resulted from episodic ocean formation driven by occasional periods of intense volcanism (Baker *et al.*, 1991). At such times, which perhaps occurred throughout much of the planet's history, atmospheric CO_2 levels are postulated to have risen well above current values. Elevated CO_2 levels and the associated greenhouse effect have also been invoked to explain the ubiquitous valley networks and higher apparent erosion rates thought to be the result of warmer and wetter conditions very early in the planet's history (e.g. Pollack *et al.*, 1987). For these epochs, high precision determination of the noble gases and their isotopes, as well as distribution and abundance of crustal volatile-bearing minerals, are the most pertinent measurements.

Nevertheless, there is at least one area where a direct connection between the current and ancient climate can be made. As Kasting (1991) pointed out, greenhouse warming from a pure CO_2 atmosphere is greatly limited in early Mars' history because CO_2 begins condensing in the atmosphere when the surface pressure reaches 300 mbar. But Kasting's calculation was based on a one-dimensional model, and it did not include the radiative effects of CO_2 ice clouds. Such clouds are believed to form in polar regions of the contemporary Martian atmosphere (Hunt, 1980; Forget *et al.*, 1996) yet very little is known about their microphysical properties, and only for pure CO_2 ice particles are we beginning to understand what their radiative properties might be (e.g. Hansen, 1996). By observing CO_2 clouds in the contemporary Martian atmosphere we can learn more about their properties and formation habits. And with simultaneous measurements of the surface pressure field, we can assess their impact on the polar energy budget. This information would provide useful constraints on the effect of CO_2 clouds on the ancient climate.

2.4. *What are the objectives of post-Viking atmospheric science investigations?*

Our current understanding of the general circulation and climate system naturally leads to a clear statement of post-Viking atmospheric science objectives. These stepwise objectives are as follows:

1. To measure and characterize the general circulation including its full global structure, and how that struc-

ture varies on time scales ranging from daily to interannual.

2. To understand the relationship between the general circulation and such aspects of its forcing as the distribution of diabatic heating, topography, surface roughness and thermophysical properties, and various external factors such as planetary size, gravity, and rotation rate.
3. To relate that understanding to the present Martian climate system, including the transport of dust and volatiles around the planet, and aeolian modification of the surface; then to extend that understanding to past climates and the evolution of the atmosphere itself.
4. To provide a quantitative basis for comparative planetary meteorology using the information gained in (1)–(3).

3. Measuring the general circulation

3.1. *How can the science objectives be achieved?*

To meet the science objectives, systematic measurements of the global three-dimensional wind field up to altitudes of 60 km or higher are required. In this section we argue that the minimum requirements include measurements of the global surface pressure field, and vertical profiles of atmospheric temperature and dust concentration. From such observations, the three-dimensional wind field, the diabatic heating rate, and the transport of heat, momentum, and dust can be calculated. This would satisfy objective (1) in the stepwise strategy outlined above for understanding the Martian climate system.

Except for sparse cloud tracking measurements, synoptic horizontal winds are difficult to measure directly on Mars, and even with clouds, the difficulty in determining their heights leads to unacceptable uncertainties in the wind level. Consequently, the best approach is to make those measurements that will allow us to construct the wind field from the dynamical equations that govern atmospheric motions. This is directly analogous to the method used to calculate winds in the Earth's middle atmosphere where temperatures and pressures are similar to those on Mars. In investigating the circulation of the Earth's middle atmosphere, it has been found necessary to combine two basic sources of data in order to calculate the three-dimensional wind field: (1) orbital remote sensing measurements of atmospheric temperature and constituent abundance as a function of height and (2) the height of a reference pressure surface derived from surface stations and/or the radiosonde network (e.g. see Crane *et al.*, 1980; Randel, 1987; Marks, 1989). The latter data set acts as a boundary condition in the calculation of winds and cannot be determined with sufficient accuracy by remote sounding from orbit.

The set of equations which need to be solved are those which govern atmospheric dynamics: Newton's second law for air flow in the rotating frame of the planet, the conservation of mass (or continuity equation), and the conservation of energy (or the first law of thermodynamics). Newton's second law when expanded in spherical coordinates leads to a momentum equation for each velocity component: zonal, meridional, and vertical. For atmospheric motions of a horizontal scale larger than several tens of kilometers, the latter can be approximated by the hydrostatic equation. When integrated this gives

$$p(\lambda,\varphi,z,t) = p_s(\lambda,\varphi,z_s,t)\exp\left(-\int_{z_s}^{z}\frac{g(\lambda,\varphi,z)\,dz}{RT(\lambda,\varphi,z,t)}\right) \quad (1)$$

where p is the pressure at height z, (λ,φ) the longitude and latitude of the location, t the time, g the gravitational acceleration, T the temperature at height z, R the gas constant, p_s the surface pressure at an altitude $z_s = 0$. Because winds fundamentally arise from pressure gradients, the pressure field at a given altitude is the essential data that allows us to calculate the horizontal air flow at that height. Clearly, to solve equation (1) and provide a pressure surface for calculating gradients, one needs measurements of surface pressure, p_s, a topography reference z_s, and vertical atmospheric temperature profiles, $T(z)$. From surface pressure data we can calculate the so-called "barotropic" component of the flow, i.e. the wind field at the surface. From temperature data we can calculate the so-called "baroclinic" component of the flow, i.e. how that wind field varies with height.[2] Therefore, the observational requirements are surface pressure, altimetry, and orbital sounding of atmospheric temperature. Furthermore, barometry and sounding must be simultaneous since a combination of surface network data taken at one time, with orbiter data taken at a different time, is not possible due to the inherently unrepeatable nature of meteorological phenomena.

In practical meteorological models and data analysis, it is common to use surfaces of geopotential

$$\Phi\left(=\int_0^z g\,dz\right)$$

and a vertical coordinate, z^*, based on the log of the pressure (e.g. Holton, 1979) where

$$z^* \equiv -H_r \ln\left(\frac{p}{p_r}\right).$$

Here $H_r(=RT_r/g)$ is a reference scale height for Mars, usually taken as 10 km, at a reference temperature T_r, and p_r is a reference pressure, typically 6.1 mbar. This coordinate system is used because g varies with latitude and altitude; it also eliminates density as a variable in the wind equations that follow. Additionally, if we use longitude and latitude as our horizontal coordinates, the so-called "primitive equations" which govern the atmosphere are (e.g. Andrews *et al.*, 1987)

$$u_t + \frac{uu_\lambda}{a\cos\varphi} + \frac{v(u\cos\varphi)_\varphi}{a\cos\varphi} - fv + \frac{\Phi_\varphi}{a\cos\varphi} + wu_{z^*} + X = 0 \quad (2)$$

[2] Note that while the barotropic component of the flow can be derived from the pressure field, the pressure field itself does not represent only the barotropic component of the flow. For example, the temporal variance of pressure may indicate the presence of baroclinic eddies.

$$v_t + \frac{uv_\lambda}{a\cos\varphi} + \frac{vv_\varphi}{a} + fu + \frac{\Phi_\varphi}{a} + \frac{u^2\tan\varphi}{a} + wv_{z^*} + Y = 0 \tag{3}$$

$$\frac{[u_\lambda + (v\cos\varphi)_\varphi]}{a\cos\varphi} + \frac{(\rho_0 w)_{z^*}}{\rho_0} = 0 \tag{4}$$

$$T_t + \frac{uT_\lambda}{a\cos\varphi} + \frac{vT_\varphi}{a} + w\left(T_{z^*} + \frac{\kappa T}{H}\right) - Q = 0. \tag{5}$$

These express, respectively, momentum balance in the zonal and meridional directions, continuity of mass, and conservation of energy. Subscripts λ, φ, z^* and t, denote partial derivatives with respect to each subscript variable. $Q = Q(T,p,\text{dust})$ represents the net radiative heating, which for Mars is strongly dependent upon the quantity of dust in the atmosphere. Other terms are as follows: (u,v,w) are the zonal, meridional and vertical wind velocities in the (λ,φ,z^*) coordinate system; X and Y are horizontal components of frictional drag per unit mass which in the free atmosphere are largely attributable to the effects of small-scale gravity waves; $\kappa = R/c_p$ where c_p is the specific heat capacity at constant pressure; $f = 2\Omega\sin\varphi$ is the Coriolis parameter where Ω is the planetary rotation rate; a is the planetary radius; and ρ_0 is the density.

Provided we know pressure surfaces as a function of geometric height, or equivalently, geopotential surfaces at log-pressure altitudes, we can proceed to derive winds, usually from approximations to the primitive equations. In the simplest analysis, the determination of horizontal winds (the zonal wind u and meridional wind v) can be done using the geostrophic approximation to the full horizontal momentum equations, i.e. $fu = -\Phi_\varphi/a$, $fv = \Phi_\lambda/a\cos\varphi$. For rapidly rotating planets, like Mars and Earth, this provides a rough estimate of the wind for large-scale and long-period flow away from the equator. However, terrestrial experience indicates that geostrophy alone can lead to significant errors ($\sim$10–30%) in the wind field (Boville, 1987). Since higher derived quantities (such as momentum fluxes) are obtained by repeated differentiation of wind fields, these will be even less accurate. A better method is to use geostrophy only as a first guess and then to implement a coupled iterative solution of "balance wind" equations, which are equations (2) and (3) with the time tendency, frictional drag and vertical advection terms omitted, but the nonlinear curvature terms retained (e.g. see Randel, 1987). Near the equator ($<20°$), where geostrophy becomes less valid, the winds can be interpolated. We note that frictional drag terms may become significant above 40 km on Mars where intermediate scale internal gravity waves and tides are predicted to break (Zurek, 1976; Barnes, 1990; Joshi *et al.*, 1995b). In this case, it may be possible to diagnose these effects as residuals.

We can also investigate the circulation in the meridional plane on Mars, i.e. the vertical winds, w. Usually, we cannot use the continuity equation (4) on its own to estimate meaningful values of w because we would rely upon horizontal derivatives of u and v which are an order of magnitude smaller than the winds themselves: a 10% error in the horizontal wind could lead to 100% error in vertical velocity. Drawing upon experience in the terrestrial stratosphere, and work done by Santee and Crisp (1995) for Mars, the vertical velocity is best obtained from the thermodynamic equation (5). To solve this equation, however, one needs to calculate the net radiative heating term, Q, which depends on the atmospheric temperature and the effect of radiatively-active constituents. On Mars, dust concentration is a basic variable that determines atmospheric heating or cooling. We therefore require measurements of vertical profiles of dust abundance in the atmosphere, and knowledge of the optical and radiative properties of the dust. Such data can be obtained from orbit using remote sounding techniques. It is likely that even if orbital sounding and surface pressure measurements are coordinated, individual grid point values of vertical velocity, w, would still be poor due to lack of surface coverage. Nevertheless, a zonal mean vertical velocity in the meridional plane would show direct observations of Mars' diabatic circulation (i.e. that driven by both eddies and radiative processes).

To investigate the general circulation of Mars, we therefore conclude that horizontal wind computation requires known elevation of surface stations, surface station pressures, and atmospheric temperatures. Vertical motion, important for the meridional circulation, additionally requires knowledge of the vertical dust distribution, and the radiative and optical properties of the dust. These measurements need to be simultaneous, global and long term for characterization of seasonal or interannual variability. In the most sophisticated analysis, a GCM incorporating a data assimilation scheme could calculate evolving wind fields from the primitive equations using continually updated data for the surface pressure field $z^*(p = p_s)$ and orbiter temperature and dust data (e.g. Lewis and Read, 1995; Lewis *et al.*, 1996).

3.2. *How do currently planned missions meet these requirements?*

Currently planned missions to Mars and their principal science objectives are listed in Table 2. NASA's Mars Surveyor Program will launch two spacecraft to Mars at every opportunity beginning in 1996 and continuing well into the next millennium. If successful, the 1996 opportunity should provide a first step towards characterizing the general circulation. NASA's Mars Global Surveyor (MGS) will aerobrake into a near-polar sun synchronous orbit with a period of just under 2 h. Though instrumented mainly for surface mapping, MGS's Thermal Emission Spectrometer (TES) will provide daily global coverage of thermal structure and dust loading of the atmosphere for one Mars year (Christensen *et al.*, 1992). On the surface, meteorological measurements will be conducted from NASA's Pathfinder lander, and the two small stations and (possibly) two penetrators flying on Russia's Mars 96 mission. All of the landers planned for 1996 will be going into the northern hemisphere.

In 1998, NASA will launch another orbiter and lander, and the Japanese will launch their Planet B aeronomy

Table 2. Currently planned missions to Mars and their principal science objectives

Launch	Mission	Objective
1996	Pathfinder	Engineering demo, *in situ* science
1996	Mars Global Surveyor	Surface mapping
1996	Mars 96	Atmosphere, surface, and subsurface
1998	98 Orbiter	Atmospheric mapping
1998	98 Lander	Volatiles and climate, polar terrains
1998	Planet B	Aeronomy
2001	01 Orbiter	Geochemistry, near surface water
2001	01 Lander	Ancient climate, prepare for SR
2003	InterMarsNet???	Network science and geochemistry
2003	TBD	TBD
2005	Sample Return (SR)	Sample return

mission. The NASA orbiter will carry a dedicated atmospheric sounder and a wide angle camera. The sounder is a limb-scanning pressure modulator infrared radiometer (PMIRR) and will conduct measurements of the thermal structure, dust loading, and water content of the Martian atmosphere from an orbit similar to that planned for MGS (McCleese *et al.*, 1992). PMIRR represents a significant improvement over TES in that it will acquire data at higher vertical resolution (1/2 scale height vs. 1 scale height), over a greater depth (0–80 km), and it will provide direct information on how dust and water vapor are distributed vertically. PMIRR, therefore, will provide an excellent definition of the baroclinic component of the general circulation. Unfortunately, only a single lander is flying in 98 so the barotropic component will not be determined. However, the lander will be going to a high southern latitude ($\sim 70°$S) and will therefore provide the first landed meteorological measurements in the southern hemisphere.

In 2001, NASA will fly the last of the original Mars Observer payload—a Gamma Ray Spectrometer (GRS) which will map near-surface elemental abundances and search for subsurface water (Boynton *et al.*, 1992). Except possibly for a camera and a relay, no other instrumentation is being considered for the 2001 orbiter because of mass constraints imposed by the launch vehicle, and the unfavorable approach geometry of this opportunity. The 2001 lander is not yet defined, but is likely to carry some kind of meteorological instrumentation. At present, therefore, this opportunity will contribute very little to general circulation science objectives.

Though we will learn much from these missions regarding the general circulation and climate system, none have the kind of coordinated simultaneous measurements from the surface and orbit that will allow us to unambiguously determine winds. At this time, the best opportunity for these kinds of measurements appears to be in 2003 where two possible scenarios can be envisioned. First, if the joint ESA/NASA InterMarsNet mission is selected, then its three quite-capable landers could be augmented with the simpler μ-Met probes and use the resources earmarked for the InterMarsNet landers to build a dedicated orbiter. In either case, *true* global network science could be realized. In the next section, we give the rationale for a global network of pressure sensors.

4. Rationale for a network only measuring surface pressure

As we argue below, 15–20 landers are minimally required to meet the science objectives pertaining to the general circulation. Yet it is not possible to fly this many landers of the type being considered in the Surveyor Program and InterMarsNet. The reason is cost. The Surveyor and InterMarsNet landers are designed to address a variety of disciplines and are therefore large, heavy, and complex—features which inevitably drive up the cost. Therefore, in order to reduce cost and still be able to establish a global network, new landers must be designed that are light, simple, and small.

This can be achieved by limiting the payload to a single discipline and, in our approach, a single measurement—pressure. In effect, we are trading the ability to make many measurements at a single site, for a single measurement at many sites—a trade we are willing to make so that 15–20 landers (or more) can be delivered for global meteorological science. Fortunately for meteorology, pressure measurements place the smallest demands on spacecraft design. Pressure sensors are light, have minimal power requirements, and do not require orientation or deployment. It is also fortunate that pressure is the most important surface meteorological measurement for enabling an accurate determination of the seasonally-varying wind systems (see Section 3). As shown in the paper by Merrihew *et al.* (1996), the μ-Met concept is feasible from an engineering standpoint: the required number of stations can be delivered and they can last a full Martian year to study seasonal variability.

5. Measurement requirements

The primary scientific objective of the μ-Met mission is to collect data that will allow us to characterize the general circulation. We have argued that this requires coordinated surface and orbital measurements. The next step is to assess the required range, precision, accuracy, and spatial extent of these measurements as they relate to the science that can be achieved. In this section, we make this assessment bearing in mind the need for consistency between surface and orbital specifications.

5.1. *Surface network*

Atmospheric waves play a key role in the atmospheric circulation on Mars and the transport of heat, momentum and constituents. Consequently, the temporal and spatial variability of atmospheric waves are cardinal factors in determining specifications relating to a network of

Table 3. An estimate of the minimum number of surface stations, N, and their distribution required to study various atmospheric phenomena independently. This table is meant to provide a rough guideline. Accounting for overlap between stations that are suitably located to gather data for more than one atmospheric feature, about 15–20 stations are minimally required to address the global climate system

Atmospheric feature—related issues	N	Location
Dynamical features		
•Zonal-mean circulation—momentum, wave processes, energy transfer	15	Spread in longitude, over $\pm 70°$ latitude
•Midlatitude waves (e.g. baroclinic and stationary)—energy transfer, dust raising	12	7 equi-spaced in the northern hemisphere around 60°N, 5 in the southern hemisphere around 50–60°S
•Equatorial waves—tracer transport, dust raising	3	15°S, equator, 15°N, at widely spaced longitudes
•Thermal forcing	—	All of above
•Mesoscale system—regional winds, frontal structures, slope winds	3	At the vertices of a triangle, with ~ 300 km sidelength, to obtain the geostrophic wind vector
General climate features		
•CO_2 cycle and mean global pressure—interaction with the general circulation, polar heat balance	16	Spread in latitude and longitude, approximately evenly. The large number is required to define the global mean pressure
•Dust storms—control of climate and aeolian features	8	2 at the equator, 2 at 35°S, 2 at 20°S, 2 at 15°N, at storm longitudes

stations. The required number and distribution of stations depends on horizontal atmospheric oscillations, and the necessary precision and accuracy of pressure measurements depends on the amplitude of wave perturbations at the ground. Because of the scattered nature of existing data for the Martian atmosphere, GCMs and other models are useful tools for addressing these issues.

5.1.1. *Number and siting of surface stations.* Our ultimate aim is that stated previously, namely to characterize and understand the general circulation. We argue that the deployment strategy for a network to study the general circulation should be prioritized according to the following spatial requirements: (1) latitudinal coverage, (2) longitudinal coverage, and (3) sites of special meteorological interest. Table 3 summarizes stations required to characterize them independently. Stations overlap in their different roles and this leads to the conclusion that about 15–20 stations are minimally needed to cover all circulation components.

Temperature, wind, water vapor, dust loading, and pressure are essentially functions of latitude. Because of this, the Martian general circulation can be divided into latitudinal bands analogous to the Earth's (Lorenz, 1967): the tropical circulation, the midlatitude circulation, and polar circulation. Consequently, to measure the general circulation primarily necessitates latitudinal deployment of stations. These would be separated by no more than 20° of latitude (the latitudinal size of characteristic weather regimes, $\sim 10^3$ km) and the distribution would extend up to polar regions $\pm 70°$ latitude. Clearly, three latitudinally-separated stations in each hemisphere and one at the equator, making seven in total, is the bare minimum of latitudinal coverage for a network that aims to study the characteristic weather regimes of the global circulation. However, longitudinal coverage is indispensable for studies of atmospheric waves that have lengths extending over many longitudes and for meaningful calculation of the zonally-averaged circulation.

Viking lander data and model results suggest that baroclinic waves in the northern hemisphere are zonally coherent waves propagating at varying phase speeds (Barnes, 1980; Murphy *et al.*, 1990; Barnes *et al.*, 1993). Assuming zonal coherence, basic harmonic analysis allows us to quantify the number of surface stations that could, in principle, unambiguously diagnose the spatial structure of a wave with a particular wave number from its pressure signature around a latitude circle. A wave can be represented by a zonal mean value plus sinusoidal terms containing phase and amplitude information. In the simplest case of wave number $m = 1$, a sinusoidal pressure wave at a certain latitude, with phase α, can be represented spatially by $p_1(\lambda) = \bar{p}_1 + A_1 \sin(\lambda + \alpha)$. The three unknowns $\bar{p}_1$, A_1 and α, necessitate a minimum of three measurements of pressure around the latitude circle to derive the mean, amplitude and phase information. More generally, for a wave number m, a minimum of $2m+1$ longitudinally-distributed stations is required to diagnose the spatial structure of the wave for that wave number. Without additional stations, wave numbers of higher order than m would represent a cut-off, or "Nyquist wave number" limit. Nevertheless, spectral analysis of time series data would still allow the existence of traveling waves with higher wave numbers to be inferred from their frequency components.

Further difficulty is encountered if we wish to estimate values of the surface pressure between stations by interpolation. Studies of the distribution of surface stations for measuring terrestrial weather fields suggest that to approximate spatial derivatives even crudely, at least five equally-spaced samples per wavelength are required for the smallest wavelength feature (see, e.g. Barnes, 1994). We conclude that the desired longitudinal separation of stations is primarily determined by the highest wave number (shortest wavelength) feature that we wish to characterize. Fortunately, because Mars is about half the diameter of the Earth and has a dynamically similar atmosphere, the dominant wave number for midlatitude baroclinic waves is correspondingly less. Spectral analysis of the Viking lander time series pressure data and output from GCMs show that the typical northern midlatitude zonal wave number is $m \leq 3$ compared to $m \sim 6$ for the Earth. This would imply that we require a minimum of seven stations approximately equally spaced around a circle of latitude in the northern hemisphere to diagnose the spatial structure of waves up to $m = 3$. Stations would be equi-spaced around 50–60°N latitude (the latitude where modeled weather variances are maximal) and, to the extent possible, in predicted storm zones where the baroclinic waves have their largest surface pressure amplitudes. In the southern hemisphere, model simulations suggest that baroclinic waves are suppressed by the effects of topography. Five stations would allow us to diagnose the barotropic component of wave number 2 planetary waves.

Banfield *et al.* (1996) have analyzed Viking IRTM (Infra Red Thermal Mapper) data to calculate weather correlation length scales at an altitude of about 25 km from temperature variances. For the two equinoxes, they found horizontal length scales (representative of transient phenomena) of $\sim$1000 km. However, scales varied from 240 to 2500 km depending on latitude and season, with a general shortening in the northern hemisphere. If extrapolated to the surface, they suggest that their results indicate that a larger number of landers is required than that inferred from Viking lander measurements, perhaps as many as 100 for the lower correlation length scale. They base this estimate on terrestrial numerical weather forecasting, where the accuracy in estimating an atmospheric state has been shown to be directly related to the ratio of the separation of surface stations to the weather correlation length scale. In contrast, Lewis *et al.* (1996), have used a data assimilation model for Mars and shown that comparatively few pressure measurements (e.g. 12 in the southern hemisphere) are significantly beneficial in reducing errors in the zonal mean wind. A highly questionable assumption of Banfield *et al.* is that temperature variance fields at $\sim$25 km are representative of surface fields. GCM results frequently do not show a strong coherence (Barnes *et al.*, 1993, 1996), though we note there is insufficient temporal and spatial coverage in the data to draw definitive conclusions. It is also important to note that the μ-Met network is not intended for weather forecasting but to diagnose the general circulation.

In addition to basic latitudinal and longitudinal coverage, sites of special meteorological interest may play a role in landing site selection. Locations where the near-surface winds are predicted or known to be particularly strong are places where surface measurements are especially useful. One location is a longitudinal band at 50°W near the equator where the cross-equatorial Hadley cell is predicted to be funneled into a low-level jet (Joshi *et al.*, 1995a). While these low-level winds could not be resolved explicitly, the combination of surface pressure data and temperature/dust profiles would allow us, in principle, to gauge the large-scale meridional circulation within a broader longitudinal band. Surface pressure, in particular, can be significantly affected by mass re-distribution due to the Hadley cell overturning (Pollack *et al.*, 1993). Other places of special interest include dust storm areas. Local dust storms are known to originate during all seasons at latitudes 10–20°S and 20–40°N so that it would be useful to place stations within these areas to quantify associated meteorological fields more accurately.

For the CO_2 cycle, the meteorological component of the annual pressure cycle (see Hourdin *et al.*, 1993; Pollack *et al.*, 1993) is smallest near the equator and therefore low latitude pressure measurements should best represent the atmospheric mass. However, since we cannot ascertain a priori how representative a single measurement will be, it is important to have many well-distributed stations to separate the weather component signal from the one due to the seasonal CO_2 cycle at all latitudes. Since the meteorological component is out of phase in the two hemispheres but the CO_2 cycle is in phase, good latitudinal coverage in both hemispheres is the primary requirement. Because the dynamical part of the pressure signal is dominated by the Hadley circulation at low and midlatitudes, much can be learned about the mean meridional circulation solely from latitudinally-distributed pressure measurements (Pollack *et al.*, 1993).

We can estimate the number of stations required to characterize the CO_2 cycle by making use of models. In Fig. 3 we show the r.m.s. error in estimating the seasonal variation of global mean surface pressure as a function of the number of stations. The "data" were taken from an annual simulation with the Ames GCM. For simplicity, we assumed simple rectangular grids and populated them in a systematic fashion (e.g. north to south, then west to east). The results for two grids are shown. Other choices give similar results. This analysis suggests that even to characterize the CO_2 cycle, a considerable number of landers is required: the r.m.s. error does not stop significantly fluctuating until the number of landers exceeds about 16 or so.

In summary, the combination of the necessary latitudinal and longitudinal coverage (regardless of additional sites) leads to the inescapable conclusion that a large number (15–20) of surface stations is required to investigate the general circulation on Mars. Networks which are smaller than this, or are not distributed for the spatial coverage described above, cannot claim to address the objective of characterizing the general circulation. This is not to say that smaller networks lack utility; on the contrary, small networks can usefully characterize aspects of the circulation and even provide information on planetary-scale phenomena (e.g. the CO_2 cycle and baroclinic waves) as demonstrated by the Viking lander data. However, for a systematic *ab initio* study of the global climate on Mars, a large network is indispensable. To put matters into perspective, the proposed number of Martian surface stations is extremely modest when we compare the terrestrial surface observation network: $\sim$2500 surface stations, $\sim$1500 ships, $\sim$500 radiosondes (weather balloons), and over 2000 daily reports from aircraft. Figure 4 shows a 16 station μ-Met network on a topographic map of Mars—this is just one possibility amongst many but provides a useful illustration of what is realistically achievable. It is important to note that the spacecraft deployment technique restricts the landing site configuration. Firstly, sites at >10 km elevation above the reference geoid are inaccessible because sufficient deceleration is not possible; generally sites <5 km elevation are preferred for reduced risk. Secondly, for the low-cost, μ-Met mission concept discussed by Merrihew *et al.* (1996), a centrifugal release of μ-Met stations leads to

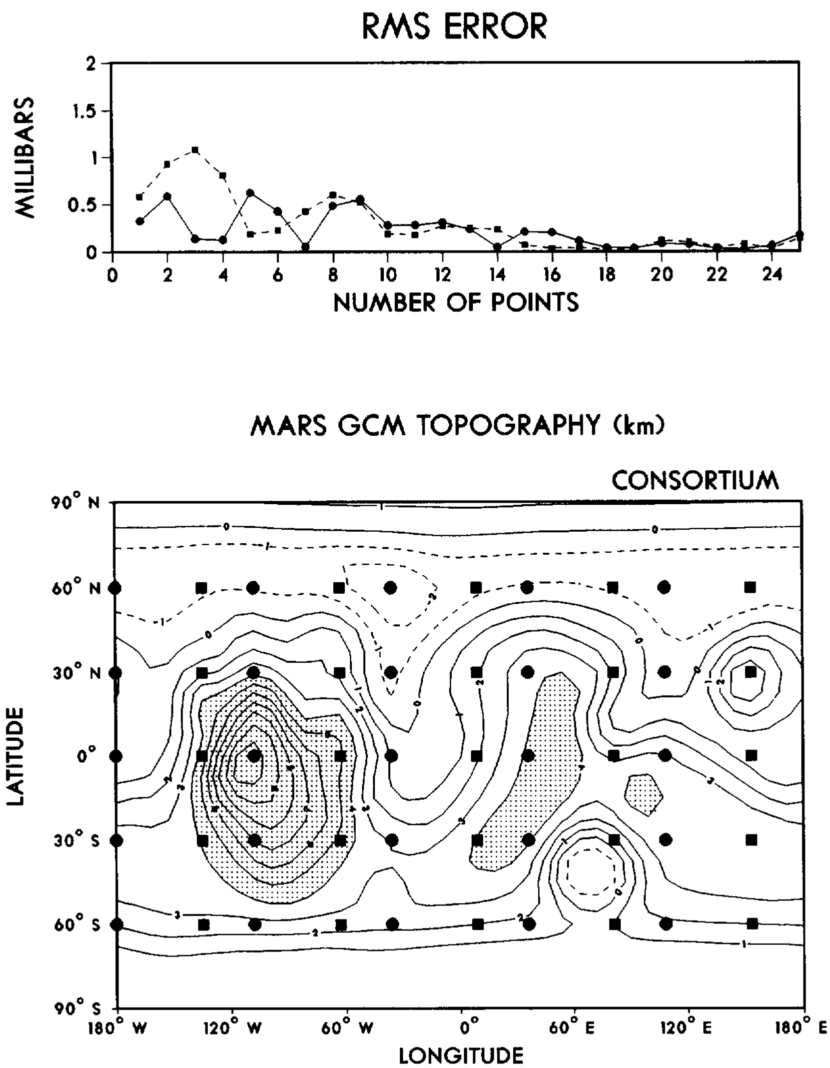

Fig. 3. Upper panel: the r.m.s. error in estimating the global mean surface pressure as a function of the number of stations using surface pressures taken from the NASA–Ames Mars GCM. The stations were distributed according to the grids shown in the lower panel. Note that 3–4 of the populated longitudes would satisfy the requirements listed in Table 3 as well as provide an accurate measure of the global atmospheric mass budget

stations lying along a locus on the planet, where the shape of the locus is determined by the carrier approach velocity and the declination of the hyperbolic approach asymptote. Figure 4 shows two such loci, sinusoidal in shape. Stations along each locus are marked by squares and circles, respectively. The two loci are produced by releasing two sets of four stations (a group of eight probes in all) for one locus, propulsively adjusting the carrier spacecraft, and releasing another two sets of four stations. The second group of eight stations actually arrives on Mars at an earlier time than the group released first due to the propulsive acceleration. This acceleration is a predetermined amount such that Mars has turned 90 deg (the phase separation of the two sinusoids) by the time the first release group of eight stations subsequently arrive. Even with these restrictions, it can be seen that a desirable distribution is achieved from the standpoint of meteorological science. There are five midlatitude stations per hemisphere (with some in the northern storm tracks), three stations in the equatorial region, and three in the subtropics, all with good longitudinal separation. Nearly all of the demands of Table 3 are satisfied.

5.1.2. *Precision, accuracy and zero-drift requirements for pressure measurement.* The barometer for each Viking lander was a commercially available type (P4A) manufactured by the Tavis Corporation (Mariposa, California, U.S.A.). It had a mass of 0.454 kg, drew a power of typically 0.16 W, had 0–18 mbar range, and its resolution was set by digitization to 0.088 mbar (Seiff, 1976; Mitchell, 1977). For the μ-Met mission, it is imperative that we use smaller sensors with lower mass so that we meet the broader mission design parameters. Currently, the most promising candidates are micromachined silicon devices. In addition to basic physical requirements, the performance characteristics of the pressure sensors directly relate to the meteorological science that can be achieved.

Precision. We require that the sensors for μ-Met have higher resolution than those for the Viking landers which was set by digitization to 0.088 mbar. This was too large for unambiguous discrimination of global normal modes with ~ 1.1 sol periods from diurnal tide signatures in the pressure record (Tillman, 1988). In particular, the suggested global normal modes have a meridional structure such that pressure signals associated with the modes are smaller at higher latitudes. For example, a pressure signal transient (possibly associated with a normal mode oscillation) evident at VL-1 in Viking years 1 and 2 at $L_s \sim 140$ was virtually indistinguishable from noise in the equivalent VL-2 pressure data (Tillman, 1988). This arose because of the large digitization step. With a wider distribution of landers, data which show these transients are potentially very useful because they would indicate a global temperature state (as suggested by the year-to-year repetition in the VL-1 data). If the pressure resolution is improved to 0.01 mbar in a subsequent mission, this would allow us to distinguish global normal modes in the data at both low and high latitudes. We could then define the meridional structure of global normal modes which would be aided by additional temperature data from an orbiter. Temperature profiles have been shown to strongly influence the period of the free modes (Zurek, 1988). A pressure resolution of 0.01 mbar would also allow us to monitor the passage of frontal systems which had associated pressure changes of the order of 0.1 mbar at VL-2, again barely distinguishable because of the comparable digitization step inherent in the Viking lander barometry (Tillman *et al.*, 1979). Although a resolution of 0.01 mbar (equivalent to $\sim 0.1\%$ full-scale) is demanding for a microsensor, our preliminary investigations show that this is feasible.

Accuracy. The desired absolute accuracy of the pressure sensors relates to the required accuracy in derived winds. For terrestrial numerical weather forecasting, synoptic scale wind errors of $\pm 3\ \mathrm{m\,s^{-1}}$ or less are needed (Bengtsson, 1979). For Mars, weather forecasting is not our objective, but above the boundary layer, Martian geostrophic zonal winds are typically 10–30 $\mathrm{m\,s^{-1}}$ so that $\pm 3\ \mathrm{m\,s^{-1}}$ accuracy is sufficient to determine the large-scale wind field even for relatively low wind speeds. To first order, the large-scale surface zonal wind is given by the geostrophic approximation with error Δu_g as follows:

$$\Delta u_g = \frac{1}{L}\left(\frac{RT}{f}\right)\left(\frac{\Delta p}{p}\right). \qquad (6)$$

Here L is the distance (along a meridian) over which we assume geostrophy and Δp the difference in measurements between two pressure sensors latitudinally separated by distance L. Taking $f \approx 1 \times 10^{-4}\ \mathrm{s^{-1}}$ (corresponding to mid-latitudes), a latitudinal scale $L = 1000$ km, and typical values for R and T the required accuracy for pressure measurement ($\Delta p/p$) is estimated as 0.6–0.8%, allowing for a spread in T. If the mean pressure, p, is 6 mbar, then Δp yields a value of ± 0.03 mbar absolute pressure sensor accuracy for each sensor, assuming a combination of r.m.s. errors from each. Note that the percentage error in the surface pressure will induce the same percentage error in the calculated pressure (or density) at all higher levels.

Zero drift. The long-term zero drift of the pressure sensors is specially important in an experiment that is designed to last at least a Martian year. We need to ensure that our knowledge of the CO_2 cycle is not compromised over time and that derived wind speeds and directions do not become increasingly inaccurate. Consequently, a zero drift of $< 0.1\ \mathrm{mb\,yr^{-1}}$ is desired, which would lead to geostrophic wind inaccuracies of typically $\pm 10\ \mathrm{m\,s^{-1}}$ or less, i.e. smaller than the typical wind speed. This is probably the most demanding technical specification for the pressure sensors, especially since gas-independent pressure sensors rely on the integrity of a sealed reference vacuum. Since the absolute pressure accuracy is clearly important, we suggest that the pressure sensors be zeroed in space just prior to their landing on Mars to take account of any in-flight drift. Furthermore, it would be wise to keep some identical sensors in the lab, on Earth, in a simulated Martian environment for the duration of the mission to compare zero drifts.

5.1.3. *Sampling rate strategy*. Fluctuations of pressure sensor observations, in principle, could range from < 0.1 s to many years. The high frequency (> 1 Hz), non-periodic variability is turbulence. Since this has very small amplitude, it will be automatically filtered by a proposed pressure sensor with resolution of 0.01 mbar. Moving to lower frequencies, theoretical expectations suggest we would expect a cut-off frequency in a pressure amplitude versus frequency spectrum, which would separate acoustic waves from small-scale gravity waves. This separation would occur at the Brunt–Väisälä frequency modified by the Doppler shift caused by the mean wind. For typical values, small-scale gravity waves on Mars are estimated to have periods greater than ~ 10 min. Traveling gravity waves would show up on a high precision Martian barometer as sinusoidal oscillations with periods 10–30 min exactly analogous to similar traces found on terrestrial barometers (Atkinson, 1981). Hence even if we wished to record microscale meteorological phenomena such as small-scale gravity waves, we would not need to sample more rapidly than about once per minute. However, for the study of the general circulation, we are interested in yet lower frequency phenomena. The barometric spectrum of interest ranges from weather systems that develop over several hours to the annual CO_2 cycle. Between these extremes, important phenomena include semidiurnal and diurnal tidal oscillations and baroclinic waves (as shown in Fig. 5). From experience with Viking lander data, a sampling rate of approximately once per hour, leading to 25 equally-spaced samples per sol, adequately facilitates analyses of these phenomena, and would allow identification of up to the fourth tidal harmonic. We note that there may be some minor contamination of pressure signals by traveling gravity waves (e.g. those associated with frontal systems) at the level of increments or decrements of ~ 0.01 mbar. It may not be possible to average these out because of the limited data acquisition, but they would not significantly affect analysis of lower frequency phenomena. More rapid sampling would place greater demands upon a station's power supply because of increased data acquisition and transmission.

5.1.4. *Station location and altitude*. Topography is the largest cause of spatial variance in the surface pressure on Mars. Therefore the determination of a surface station's altitude can influence the accuracy of derived winds. The horizontal location of a station does not need to be known to a high accuracy for measuring the general circulation: a 1° error in latitude or longitude (≈ 60 km at the equator)

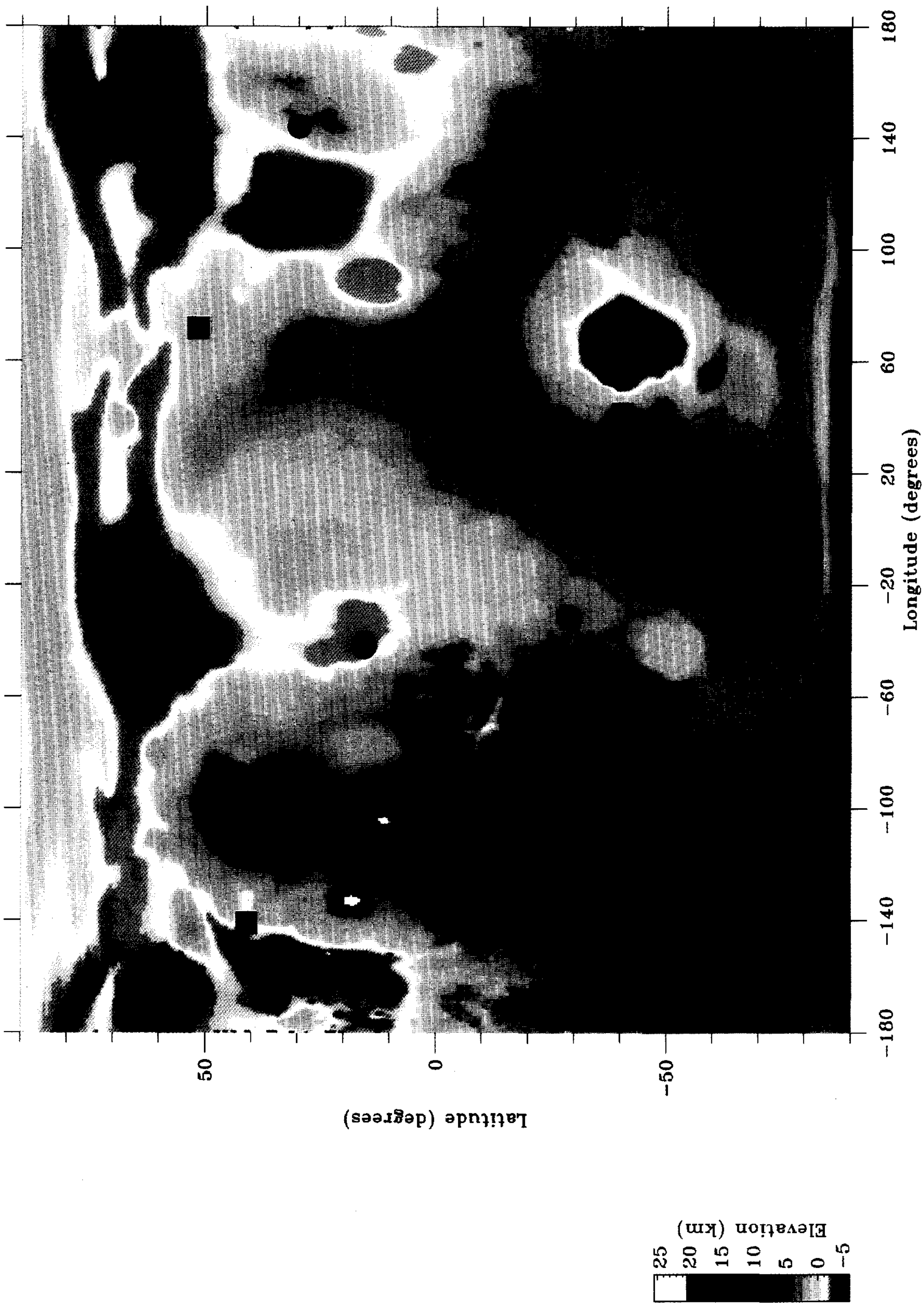

Fig. 4. An illustration of a possible network of 16 μ-Met stations on Mars. This network is achievable using centrifugal release from a spin-stabilized carrier spacecraft and a single propulsive adjustment to that carrier. Two groups of eight stations follow two sinusoidal loci and are marked by magenta squares and solid circles, repectively (see text for details). Topography is derived from the USGS digital terrain model smoothed into 600 by 300 bins

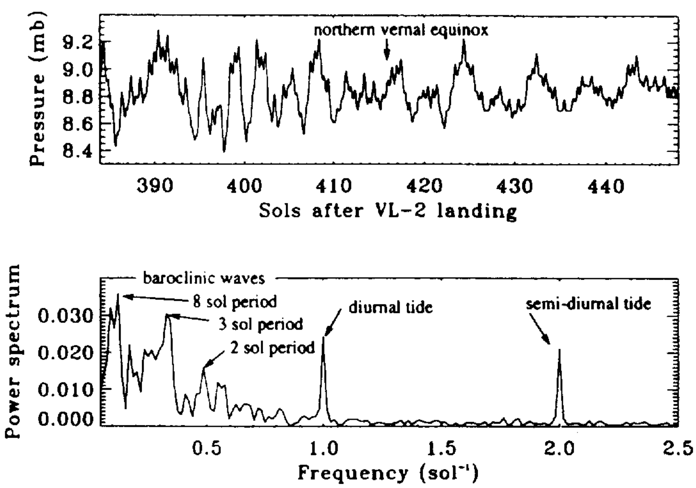

Fig. 5. Upper graph: the pressure trace at VL-2 (48°N) taken over 64 sols centered around the vernal equinox. Pressure values are in 25 bins per sol (with linearly interpolated values to fill gaps in the data) and smoothed to remove the effects of coarse digitization. Lower graph: the power spectrum of the pressure trace. This shows that baroclinic waves dominate over tides during winter/early spring at this northern midlatitude

may lead to a geostrophic wind error $\sim 5\%$. The exact coordinates of a station, however, are important insofar as they influence our knowledge of a station's altitude. The fractional pressure error resulting from an uncertainty in height Δz is given by the differential form of the hydrostatic equation, assuming Δz is small compared to the scale height:

$$\left(\frac{\Delta p}{p}\right) = -\left(\frac{g}{R\bar{T}}\right)\Delta z.$$

For typical values ($g = 3.72\,\mathrm{m\,s^{-1}}$, $R = 190\,\mathrm{J\,kg^{-1}\,K^{-1}}$ and $\bar{T} = 220\,\mathrm{K}$), the station elevation must be known to ± 70 m or better, so that the equivalent error in the derived pressure surface is not $>0.6\%$. Otherwise an error in the surface geostrophic wind of $>3\,\mathrm{m\,s^{-1}}$ results. We conclude that a knowledge of surface station elevation relative to one another to ± 70 m is a requirement for the network. We also note that relative elevation depends on our knowledge of the mean temperature in the lowest layer of the atmosphere above the surface.

The Mars Observer Laser Altimeter (MOLA) instrument on Mars Global Surveyor (MGS) will provide a global topographic map with a 0.2° (≈ 12 km at the equator) grid and a nominal ± 30 m vertical accuracy (Zuber *et al.*, 1992). In terms of limiting specifications, MOLA has a horizontal accuracy of 300 m and a local vertical resolution of 2 m. The vertical accuracy error of ± 30 m largely arises from the uncertainty in the MGS orbiter altitude. In principle, if prospective landing sites were identified early enough, the area within landing site error ellipses could be mapped at the highest horizontal and vertical resolution in advance of a future mission. When combined with Doppler telemetry data from an orbiter for μ-Met (which we estimate can give the horizontal location of a station to ± 250 m along track and ± 600 m across track), the desired elevation accuracy for surface stations could be met, if there is no unusual topographic variance within the footprint area. The latter assumption could be checked with images from the Mars Observer Camera (MOC) instrument also on MGS; these pictures have 1.5 m pixels at the highest resolution and more routine intermediate mapping has 250 m nadir pixel resolution. Although assessment of Mars landing sites is listed as an aim of MOLA (Zuber *et al.*, 1992), in practice, it is unlikely that MOLA will be allotted project time to make high resolution elevation maps of many potential landing sites for a possible future μ-Met mission. Such sites are thousands of square kilometers because of landing error ellipses and will not necessarily be of much geological interest. However, we estimate that Doppler telemetry from a μ-Met orbiter alone will only give station elevation to ± 450 m accuracy.

An alternative method is to use the annual mean pressure at each site to give an estimation of the relative station altitude. The annual average pressure for each station strongly reflects the topography but there is also a residual component due to the redistribution of the mass of the atmosphere by seasonal meteorology (Hourdin *et al.*, 1993, 1995; Pollack *et al.*, 1993). The weather component does not necessarily average to zero over a year. For given locations, GCMs can tell us the magnitude of the meteorological component in the annual or seasonal average pressure, given additional dust loading and temperature data from an atmospheric sounder. If this dynamical component is iteratively removed from measured mean pressures, we can define the relative topography between surface stations. If we accurately know the elevation of at least two sites from MOLA and MOC data, some validation would be possible. However, this technique does place some faith in the accuracy of Martian GCMs and relies on surface station longevity. Consequently, an independent estimate of each station elevation is highly desirable.

Nevertheless, as a demonstration of how one could use an annual cycle of pressure data in a simple way to derive relative station altitudes, we have estimated the elevation between the Viking landers using the measured barometric pressure at each location. Figure 6a depicts the natural logarithm of the ratio of the sol-mean pressures at the two

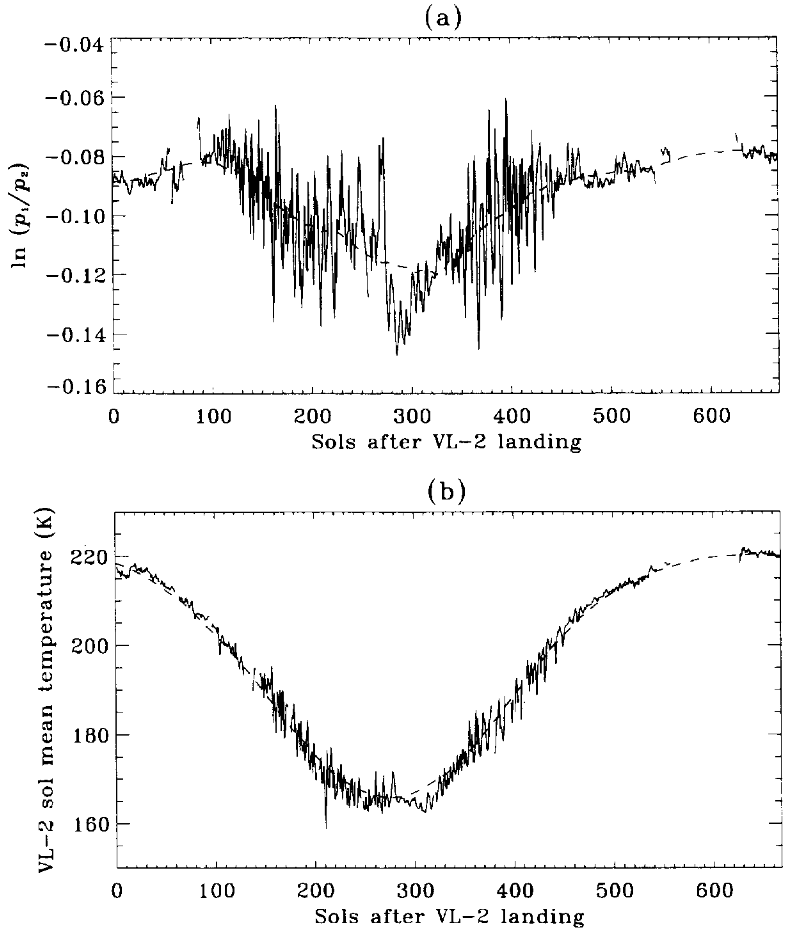

Fig. 6. (a) The natural logarithm of the ratio of the sol-mean pressures at the two Viking lander sites over a Martian year starting from when VL-2 landed. The dashed line is a smoothed version of the data using a simple box-car average of 101 sols centered on each point. A sinusoid with a period of 1 yr is evident in the data. This is primarily due to the seasonal temperature change. Sol 262 corresponds to northern winter solstice ($L_s = 270$). (b) The sol-mean temperature at the VL-2 site measured over a Martian year plotted on the same horizontal scale as in (a). The dashed line is a smoothed curve indicating the basic seasonal temperature cycle

Viking lander sites over a Martian year starting from when VL-2 landed. In principle, this is linearly related to the elevation difference, Δz, between the two landers and the mean temperature of a layer of atmosphere between them by the integrated hydrostatic equation.

$$\ln\left(\frac{p_1}{p_2}\right) = -\left(\frac{g}{R\bar{T}}\right)\Delta z \tag{7}$$

where p_1 and p_2 are the pressures at VL-1 and VL-2, respectively. Consequently, the annual sinusoidal cycle clearly evident in Fig. 6a is largely due to the annual change in the mean atmospheric temperature T. This is made obvious by a comparison of the measured air temperature cycle at VL-2 over the same period (Fig. 6b). For μ-Met stations, it is not planned to measure the *in situ* temperature, but temperature data would be produced from orbital sounding. Representative annual temperature cycles could be deduced from the orbital data for given locations with the aid of models. If we take the sol-mean temperature at VL-2 as representative of the mean temperature of the thin layer of atmosphere above VL-2 (some $\sim$900 m up to the VL-1 altitude level from a first estimate of the elevation difference) then we can eliminate the temperature-dependent annual cycle in Fig. 6a. The data were first spline-filled to take account of missing values and then smoothed using a running box-car average of 101 sols centered around each point. The latter duration was selected because cycles of 10–20 sols were evident in both Fig. 6a and b so that 5–10 cycles were included for each averaged point. From the smoothed data (shown as dashed lines in Fig. 6), a derived elevation was calculated as shown in Fig. 7. Here fluctuations about the mean are a result of not taking into account the dynamical redistribution of atmospheric mass, e.g. the northward decrease of pressure due to geostrophy (Hourdin *et al.*, 1993). Nevertheless, treating these effects as noise, an annual average from Fig. 7 gives 930 ± 45 m as the elevation difference. (If VL-2 temperatures are decremented to take account of the typical lapse rate (~ 2.5 K km^{-1}) in the thin layer of atmosphere above VL-2 this makes only a minor difference to the derived mean elevation difference: 925 ± 45 m.) Our result compares favorably with a value of 900 ± 90 m relative to the geoid (see Kieffer *et al.*, 1992) derived from lander tracking data

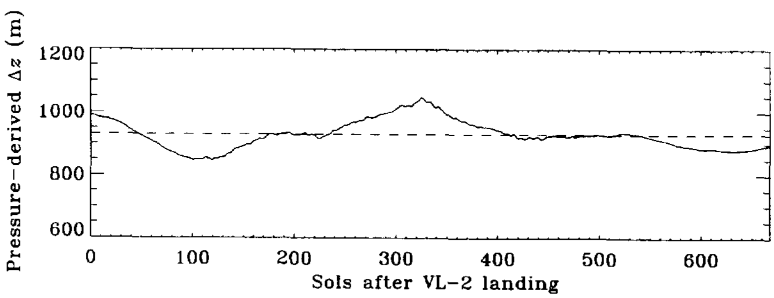

Fig. 7. The elevation difference between the two Viking landers derived from data presented in Fig. 6. The dashed horizontal line indicates the annual mean which is taken to be representative of the actual fixed difference in altitude between the two sites

(Michael, 1979) combined with the geoid defined by Wu (1981). We note that the method used here to derive the Viking lander elevation difference would not be ideal for some other locations on Mars where significant dynamical effects would need to be estimated explicitly rather than treated as noise (e.g. Hellas). However, we have demonstrated the plausibility of using annual pressure data to derive the relative elevation of stations within the accuracy specification desired for the μ-Met mission.

5.2. *Orbital remote sensing*

In Section 2, it was argued that to measure the general circulation on Mars, we require vertical profiles of atmospheric temperature and dust concentration. For sufficient global coverage, such measurements can only be accomplished from an orbital platform. Here we address some of the measurement issues as they relate to the μ-Met mission.

5.2.1. *Spatial and temporal coverage.* Orbital measurements in the μ-Met mission provide the time-varying thermal structure of the atmosphere. To produce data that can be optimally combined with surface stations, the horizontal coverage should be compatible with the distribution of surface stations, i.e. we desire observations of similar extent in longitude and latitude. In terms of orbital geometry, a low-altitude, high inclination orbit is desirable so that the satellite orbits the planet frequently and provides good longitudinal and latitudinal coverage. For example, ≥ 12 orbits per sol would give tracks separated by $\leq 30°$ in longitude at the equator. This would enable the identification of atmospheric waves up to wave number 6 or more in the retrieved temperature field, and would require an orbital altitude of ≤ 522 km for a near-circular orbit. For sufficient latitudinal coverage, the orbital inclination should be high, $\geq 70°$ to be consistent with the desired lander distribution. The combination of these factors implies that a low-altitude, circumpolar orbiter (ideally near 90° inclination) would be optimal for the characterization of the atmospheric dynamics, in particular the zonal wave numbers, phase speeds, and meridional extent of planetary waves. However, we note that tides are difficult to diagnose in this configuration because of the tendency for a polar orbiter to sample the same local time in consecutive orbits. Some analyses of the baroclinic component of tides may be possible from day–night temperature differences on a constant pressure surface or if the atmospheric sounder has the capability to view local time differences by looking in the cross-track direction. These analyses could be tied to the barotropic tidal component which would be well defined by the surface pressure network.

5.2.2. *Measurement specifications.* Temperature and dust profiles are desired from the surface up to 60 km since this is the maximum height to which dust is raised from Mariner 9 limb images (Leovy *et al.*, 1972). Because characteristic thermal structures in the Martian atmosphere are 1–4 scale heights in extent (Conrath, 1981; Barnes *et al.*, 1993; Banfield *et al.*, 1996), a vertical resolution of at least 1 scale height (~ 10 km) is desired. For example, low-altitude, vertically propagating waves have vertical wavelengths of 10–30 km (Seiff and Kirk, 1977) which could not be measured with a lower vertical resolution. In general, limb-viewing instruments are best at providing high vertical resolution.

Typical climatological thermal variability in the lowest scale height of the atmosphere is ~ 10 K per sol for a "clear" atmosphere optical depth of 0.5 (Pollack *et al.*, 1979). This would imply that the mean temperature of layers must be determined to an accuracy of better than ± 5 K merely to characterize the diurnal cycle. More importantly, let us also consider the effect of temperature sounding accuracy on the derived geostrophic wind, or its surrogate, the pressure field. The error Δp_2, in a pressure level p_2, at altitude z_2, calculated using the mean temperature $\bar{T}$ of a layer from z_1 to z_2 can be derived from vertically integrating the hydrostatic equation and differentiating with respect to temperature to give

$$\frac{\Delta p_2}{p_2} = \left(\frac{g(z_2 - z_1)}{R\bar{T}^2}\right)\Delta \bar{T}$$

where $\Delta \bar{T}$ is the temperature uncertainty. To be consistent with our desire to characterize large-scale winds in the lower atmosphere to $\pm 3\ \mathrm{m\,s^{-1}}$, the mean temperature of a layer of ~ 10 km vertical extent should be determined to an accuracy of better than ± 2 K. Also, as atmospheric layers are integrated upwards, temperature errors will have a cumulative effect so higher level winds are likely to be more inaccurate. This just reflects the basic dynamics: the wind shear is determined by the three-dimensional temperature structure whereas the surface geostrophic wind is a function of the horizontal surface pressure gradient. An accuracy of better than ± 2 K is also required to diagnose thermal signatures of transient weather activity

Table 4. Basic science specifications for the μ-Met mission. It is assumed that 15–20 landers would be instrumented as specified and there would be a single orbiter. The orbiter would also act as a communications relay for the landers

Parameter	Specification
Lander	
Instrument	Miniature pressure sensor
Pressure range (mbar)	0–12
Pressure accuracy (mbar)	0.03
Pressure precision (mbar)	0.01
Long-term drift (mbar yr^{-1})	<0.1
Lander relative elevation accuracy (m)	70
Pressure samples per sol	25
Orbiter	
Instrument	Atmospheric sounder
Temperature accuracy (K)	≤2
Vertical resolution (km)	≤10
Orbit altitude (km)	≤520
Orbital inclination (°)	≥70

which have variances of between 1 and 10 K^2 depending on latitude, season and altitude (Barnes *et al.*, 1993; Banfield *et al.*, 1996).

Because suspended dust in the atmosphere significantly affects atmospheric heating, it is required that the atmospheric sounder also conducts dust measurements. The principal parameters are the spatial and temporal distribution of dust, and dust properties for defining the radiative effects of dust in the solar and thermal parts of the spectrum. In the visible spectrum, the important factors used to determine the net heating are the visible optical depth and the particle scattering phase function. The latter is characterized in terms of the single scattering albedo (the ratio of radiation scattered to the total scattered and absorbed), the asymmetry factor (the degree of forward or backward scattering), and extinction efficiency (the ratio of interaction cross-section to geometric cross-section). In the thermal infrared, additional parameters are the wavelength-dependent emissivity of the dust and the infrared optical depth. Dust properties have been assessed from analyses of Viking lander sun diode data and Mariner 9 IRIS measurements (Kahn *et al.*, 1992); further data from the Mars Surveyor orbiters will help determine the degree to which these earlier measurements are generally representative. However, an independent measure of dust properties from a μ-Met orbiter would be highly desirable to take account of their temporal variance. Table 4 summarizes the measurement specifications discussed in this section.

6. Conclusions

In this paper, we have presented the science rationale and measurement requirements for a mission whose objective is to characterize the general circulation of Mars. This has the highest priority for post-Viking atmospheric science investigations because the general circulation controls the current climate system, and because it offers a natural laboratory for the study of Earth-like planetary atmospheres. We reviewed the outstanding questions regarding the general circulation on Mars and these lead to a set of stepwise science objectives: (1) to measure the general circulation, (2) to understand how it is forced, (3) to relate that understanding to the climate system, and (4) to provide a quantitative basis for intercomparison of planetary atmospheres.

Although many Mars missions are planned for the next decade and will address some aspects of these objectives, none have the kind of coordinated measurements that are needed for the full achievement even of objective (1). For this, we need to unambiguously determine winds, globally and for a full Martian year. The well-known equations which govern atmospheric circulation suggest the required observations: surface pressure measurements from a globally distributed network of landers and simultaneous orbital remote sensing measurements of the thermal structure and dust loading of the atmosphere. The surface network provides the horizontally varying component of the flow, while the remote sensing data gives the vertically varying component. Both components are needed to construct the wind field. Furthermore, we have argued that at least 15–20 landers are needed to adequately map the pressure field, and that a high-inclination short-period orbiter is needed for the remote sensing measurements. A large number of landers is required because the important components of the general circulation have pronounced latitudinal and longitudinal dependencies; a high-inclination short-period orbiter is necessary for daily global coverage.

An important part of the rationale for a network of landers measuring only pressure is that it leads to the design of probes that are sufficiently small and light that the required number can be delivered by just a single, low-cost launch vehicle (see the paper by Merrihew *et al.* (1996)). Pressure sensors do not require orientation or deployment and place minimal demands on spacecraft resources. In particular, the small power demand is of critical importance in a mission which must last a Martian year (1.9 Earth years) to cover all seasons. However, to satisfy the science objectives, we have shown that the pressure sensors must have high precision (0.01 mbar) and accuracy (0.03 mbar) with minimal long-term drift (<0.1 mb yr^{-1}). These are demanding, but achievable, specifications. We also need to know station-to-station elevation differences to within ± 70 m; we have presented a plausible method to achieve this using the pressure data itself should more direct methods be unable to meet this requirement. Given these specifications, and the ability to retrieve temperature profiles to ± 2 K with ≤ 10 km vertical resolution, we can determine the surface wind field (above the boundary layer) to ± 3 m s^{-1}, an accuracy commensurate with terrestrial experience.

Finally, we suggest that the best opportunity for carrying out the μ-Met mission is in 2003, either to augment the joint ESA/NASA InterMarsNet mission if it is selected, or as a NASA-only mission if it is not. If InterMarsNet is selected, then the ESA-supplied orbiter could provide the required remote sensing measurements. If InterMarsNet is not selected, then NASA could use the resources allocated for the InterMarsNet landers to build a dedicated orbiter. In either scenario we would be able to construct, for the first time, the global time-varying wind field from another planet. Although this is grand, large-scale science, we have shown how it can be accomplished with a relatively small-scale, low-cost mission.

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